Influence of rain re-evaporation on Pacific rainfall patterns in an AGCM

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Abstract. Sensitivity experiments with an AGCM show that parameterized rain re-evaporation has a large impact on simulated precipitation patterns in the tropical Pacific. Specifying strong re-evaporation rates in the model leads to a better overall agreement of simulated precipitation with observations. Weak re-evaporation leads to the formation of a "double ITCZ" during the northern warm season, and an insufficiently vigorous south Pacific convergence zone (SPCZ) during the northern cold season. Analysis of water vapor budgets suggest that this sensitivity of may be related to differences in free-tropospheric water vapor transport in the ITCZ/SPCZ complex of the central and western Pacific. An idealized AGCM experiment incorporating a fictitious drag source in the tropics provides further evidence to support the role of water vapor transport in suppressing the double ITCZ bias.

1. Introduction and Background

For most of the year the monthly-average precipitation over the Pacific ocean is dominated by a single band of intense rain rates centered between 8N and 10N, the so-called intertropical convergence zone (ITCZ). The dynamical reasons for this asymmetry in nature are not known. In atmospheric global circulation models (AGCMs) good simulations of the mean precipitation over the tropical Pacific are difficult to obtain. A frequent bias in AGCMs is the formation of a spurious second ITCZ in the southern hemisphere (e.g.; Meehl and Arblaster, 1998). While nature does show hints of a southern ITCZ over the Pacific, particularly during March through May (Zhang, 2001), this feature in AGCMs is usually too strong and persistent, lasting through the northern warm season June-September. The occurrence of double ITCZs in AGCMs leads to large rms errors in simulated precipitation, since it represents spurious rearrangement of the most intense precipitation on earth. Connections between double ITCZs and other AGCM simulation biases have not been conclusively established. However, it is clearly of concern to climate modelers, if AGCMs are producing large errors in the horizontal distribution of atmospheric latent heating. Finally, the wide distribution and similar structure of this bias in a variety of AGCMs suggests a the existence of a shared misunderstanding in current implementations of convection parameterizations.

Other researchers have examined mechanisms that control the location of ITCZs in the tropics. Chao (2000) and Chao and Chen (2001) propose a competition bewteen rotational deflection of convergent winds and wind driven surface evaporation @?? to explain the location of ITCZs close to, but not on, the Equator. As f increases away from the Equator, Coriolis effects can induce larger surface wind perturbations which in turn lead to increased surface evaporation, but at the same time, low-level convergence induced by heating becomes weaker. The role of high frequency transient motions in determining the location of the ITCZ has been studied extensively (e.g.;

Holton et al., 1971; Lindzen, 1974; Hess et al., 1993). Hess et al. showed that the character of both transients and precipitation in a zonally and equatorially symmetric aquaplanet depended strongly on the convective parameterization employed. With moist convective adjustment a single equatorial ITCZ formed, coupled with weak, poorly organized easterly wave activity. With a CAPE-based, mass flux scheme double off-equatorial ITCZs formed near 10°S and 10°N, along with intensified easterly wave activity. Philander et al. (1996) and Li (1997) examine the origin of the observed north-south asymmetry in ITCZ location using coupled ocean/atmosphere models. They find that weak initial asymmetries in atmospheric circulation arising from factors such as continental distribution, can be amplified by air-sea interactions, which favor one ITCZ over the other. Recent AGCM simulations over a "swamp"-planet, i.e., a mixed layer driven by surface fluxes only, obtain a single, but relatively diffuse, equatorial ITCZ (M.-I. Lee, personal comm.).

This study will not focus on the fundamental processes which determine ITCZ location in idealized settings. The simulations conducted here are over observed, time-varying, SST distributions (Reynolds, 1988). The AGCM used in this study (NSIPP-2.0) employs a CAPE-based, mass-flux convection scheme - Relaxed Arakawa-Schubert (RAS) (Moorthi and Suarez 1992). We expect that this model would produce off-equatorial, double ITCZs given equatorially symmetric, fixed SST forcing as in Hess at al. (1993). Nevertheless, when forced by observed SSTs, NSIPP-2.0 exhibits an interesting sensitivity in ITCZ structure to the strength of parameterized rain re-evaporation. As re-evaporation increases, the tendency for the model to form a spurious southern ITCZ decreases. The previous version of the NSIPP AGCM (NSIPP-1) (Bacmeister and Suarez 2002) exhibited similar sensitivity. However, separate treatments of re-evaporation for large-scale and convective rain in NSIPP-1 complicated the interpretation of the results (unpublished). Although this sensitivity has been useful in empirical "tuning" of the NSIPP AGCM to optimize precipitation simulations, the physical origin of the sensitivity was not understood. Unfortunately, anecdotal evidence from other modeling groups suggests that this straightforward sensitivity is not universal (I. M. Held, personal comm.), and also that other sensitivities may exist to parameters such as cumulus friction. In this study we use NSIPP-2.0, which employs a single parameterization of re-evaporation, in an attempt to isolate to the processes connecting rain re-evaporation to the seasonal mean rainfall patterns in the tropical Pacific. We hope that this will shed light on the general problem of tropical precipitation modeling, and suggest reasons to explain the variety of parameter sensitivities exhibited in AGCM rainfall simulations.

The paper is organized as follows. Section 2 provides a description of the AGCM used in this study. The parameterization of rain re-evaporation is described in some detail. Section 3 outlines the AGCM experiments performed. Section 4 presents the basic sensitivity of the model simulations to re-evaporation. Seasonal mean fields are shown, as well as some analysis of high frequency transients, thermodynamic vertical profiles, and simulated atmospheric re-evaporation rates. Section 5 describes an analysis of the water vapor budget in domains around the southern and northern Pacific ITCZs. This analysis reveals interesting differences in free-tropospheric water vapor transport that depend on rain re-evaporation. Section 6 discusses results from an experiment with added drag in the tropical free troposphere. This experiment is intended to isolate the role of transport in the maintenance of the southern ITCZ. The behavior of high frequency transients in our simulations is discussed in Section 7.

2. Model Description

We use a preliminary version of the NSIPP-2 AGCM (NSIPP-2.0) for this study. NSIPP-2.0 was developed from the NSIPP-1 AGCM, which was documented in

Bacmeister and Suarez (2000) and Bacmeister and Suarez (2002). Simulated seasonal means and responses to interannual SST variation in NSIPP-1 were both in good agreement with meteorological analyses (e.g.; Schubert et al., 2001, 2002). The significant modifications to NSIPP-2.0 and NSIPP-1 involve the cloud, boundary layer, and convection schemes. These include introduction of a prognostic cloud scheme in place of the Slingo (1987)-type diagnostic scheme used in NSIPP-1, as well as a simple moist boundary layer entrainment scheme, which is called in addition to the existing first-order dry turbulence parameterization of Louis et al (1982). These modifications were aimed at improving the models simulation of subtropical marine stratus decks. Since they have little impact on the ITCZ sensitivities examined in this study, they will not be described further here.

The dynamical core of NSIPP-2.0 is the same as in NSIPP-1 and is described in Suarez and Takacs (1996). Radiative effects in NSIPP-2.0 are parameterized using the approach of Chou and Suarez (1992). Land surface effects a parameterized according to Koster and Suarez (1996), and orographic wave drag is treated according to Zhou et al. (1996).

2.1 Convection

Convection in the NSIPP AGCM is parameterized according to the relaxed Arakawa-Schubert (RAS) scheme of Moorthi and Suarez (1992). The implementation of RAS in NSIPP-2.0 is modified to include a simple cloud condensate calculation with autoconversion to rain (Section 2.3) and re-evaporation of precipitation (Section 2.2). RAS works by invoking a series of linearly-entraining plumes (or “cloud-types”) that detrain at selected levels in the vertical. Consistency is achieved by calculating the entrainment rate necessary to ensure zero buoyancy at the selected level. These entrainment rates λ are imagined to be roughly related to plume diameter according to λ=0.2/D (@Simpson 19??). Other researchers have found improved performance when lower limits are placed on λ, e.g., improved sub-seasonal variability (@Tanaka, 19??). Here we use λ_{min}=0.2/D_{BL} where D_{BL} is an estimated boundary layer depth. Our implementation invokes 10 cloud-types per gridbox per physics time-step. These are drawn at random from a uniform distribution in σ.

2.2 Rain re-evaporation

A significant conceptual improvement to NSIPP-2.0 is a more consistent treatment of large-scale and convective rain re-evaporation. In previous versions of the model, convective re-evaporation was based on microphysical expressions from Sud and Molod (1988), while large-scale rain was simply re-evaporated to grid box saturation at each time step. In NSIPP-2.0 the same formulation is used for both large-scale and convective precipitation. Microphysical expressions for rain re-evaporation are typically very complex since integrals over the Marshall-Palmer distribution (Marshall and Palmer, 1948) are performed at each step of a calculation involving droplet radii and fall speeds (e.g; Lin et al., 1983). These expressions are clearly preferable in high resolution calculations with prognostic precipitation species. However, in an AGCM it can be difficult to reconcile such expressions with the significant uncertainties in subgrid scale variability as well as with the large time steps ~1000 s typical of AGCM physics parameterizations.

We have adopted a compromise "single mode" approach, in which a representative droplet size r_p and fall speed is determined from a MP-distribution for a number of "showers" within each grid box. The preliminary version of the model used here uses only two such showers - large-scale (LS), and convective (CN). First, 3D precipitation fluxes $\mathcal{P}$ are estimated from the mixing ratios of precipitating condensate q_p according to:

$$\mathcal{P}^*_{(LS,CN)l} = \beta^{-1}_{(LS,CN)} \left(\frac{\rho_l \; \Delta z_l \; q_{p(LS,CN)\;l}}{\Delta t} \right) + \mathcal{P}_{(LS,CN)l-1} \quad (1)$$

where l represents a vertical level and $l-1$ is the level above. This assumes precipitating condensate is cleared from each grid box in a single time step. The parameter β represents the fractional area of each gridbox covered by the showers. We use β=1 for both LS and CN rain, i.e. we assume that precipitation is distributed uniformly through the gridbox. The second assumption is clearly suspect in the case of convective precipitation, but serves a useful baseline for future refinement.

Once $\mathcal{P}_{(LS,CN)}$ have been estimated, we use the third moment of the MP distribution to determine typical droplet sizes and fall speeds for the LS and CN precipitation streams. These are then used to calculate microphysically-based rain evaporation amounts,

$$\delta q_p = \alpha_r Ve(r_p) \frac{1-U}{\rho_w (A+B) r_p^2} \left(\frac{\Delta z}{w_f(r_p)} \right) q_p. \tag{2}$$

Here α_r is an empirical nondimensional parameter that modulates the strength of rain re-evaporation. U is the environmental relative humidity, assumed here to be the gridbox value. The quantity $Ve(r_p)$ is a ventilation factor (e.g.; Liou, 1992), typically between 1 and 5, that accounts for enhancements in evaporation due the air flow past the falling drop. A and B are temperature, pressure and humidity dependent microphysical quantities (e.g.; Liou, 1992; Del Genio et al., 1996) The quantity $\frac{\Delta z}{w_f(r_p)}$ is simply the amount of time spent by a drop in the layer. We assume this is shorter than the typical physics time step in our model 1800 s. For very weak showers and thick mid-tropospheric layers it is possible that this will not be true. However, no provision is currently made for this possibility other than an overall restriction on δq_p to values $\leq q_p$.

After δq_p is determined from (2) $\mathcal{P}_l$ is updated by using MAX$[q_p - \delta q_p, 0]$ in place of q_p in (1) and the calculation proceeds to the next level down $l+1$. This approach is crude, but the connections to AGCM approximations are clear, and most essential microphysics are included.

2.3 Autoconversion

The model used here still employs two distinct calculations for autoconversion. For convective rain, autoconversion is represented as a contribution to the vertical gradient of condensate in the steady-state plumes (cloud-types) invoked by RAS

$$\partial_z q_{c_{RAS}} = \ldots + (w_u \tau_{\mathcal{A}})^{-1} q_{c_{RAS}}$$

where $q_{c_{RAS}}$ is the condensate profile within an individual RAS plume, w_u is an updraft speed estimated by integrating the buoyancy force in the vertical, and $\tau_{\mathcal{A}}$ is constant time scale taken to be 1000 s. For large scale autoconversion a temperature and q_c dependent similar to that in Sud and Walker (1999) is used.

In practice, the choice of the convective autoconversion constant $\tau_{\mathcal{A}}$ has little impact on the overall precipitation simulation. It has a large impact on the partition between large scale and convective precipitation. With the value used in the experiments here, a global mean of convective to total precipitation $\frac{\mathcal{P}_{CN}}{\mathcal{P}_0}$ between 0.25 and 0.35 results. However, in the tropics most of the "large-scale" rain originates in detraining anvils. So, as long as no distinctions in the value of β in (1) are made, the distinction between large scale and convective rain is largely semantic.

3. Experimental Setup and Analysis Procedure

In this study we will discuss results from 3 baseline experiments which are identical except for the value of α_r used (Table 1) . These experiments were initialized on June 1 1989 using initial conditions from an AMIP style run using the previous version of the NSIPP AGCM, and allowed to run through December 31 1991. Horizontal resolution of 2.0×2.5 with 40 unevenly spaced σ–levels in the vertical were used for all experiments. Extensive suites of diagnostic outputs were saved from each experiment, including full 3D, daily-averaged wind, vertical motion, moisture and temperature on σ–surfaces. Most of the individual moisture tendency terms were also stored daily, including the tendency due to rain re-evaporation $\mathcal{R}$

3.1 Fictitious drag experiments

In addition to the 3 baseline experiments, we also performed a number of idealized experiments to illustrate the role played by water vapor transport in ITCZ dynamics. These were motivated by the results of water vapor budget analyses of the baseline experiments (Section 5). We will discuss the results of one of these -experiment DM, in which a fictitious drag $\mathcal{D}_m$ was introduced into in the meridional momentum equation:

$$\mathcal{D}_m = -\tau_m^{-1} G(\phi, \sigma)\ v \tag{3}$$

where the drag time scale τ_m is chosen to be ≈ 1 d. G is a function of latitude ϕ and σ given by

$$G(\phi, \sigma) = \begin{cases} \exp\left[-\left(\frac{\phi+8^\circ}{10^\circ}\right)^2\right] & \sigma \leq 0.8 \\ 0 & \sigma > 0.8 \end{cases} \tag{4}$$

This drag thus acts primarily above 800 mb in the southern tropics. The experiments performed and their shorthand designations are summarized in Table 1.

4. Basic Model Sensitivity to Re-evaporation

4.1 Mean seasonal climate

July-August (JJA) 1990-91 seasonal mean precipitation fields for experiments B1, B2, and B3 are shown in Figure 1, along with observational estimates of precipitation rates from CMAP (Xie and Arkin, 1997). The results illustrate the important climatological control exerted by the re-evaporation strength in the NSIPP AGCM. Exp B1 (Fig. 1a) tends toward a "double ITCZ" configuration, with precipitation rates in excess of 8 mm d^{-1} extending in a narrow, zonally-aligned band along 10S well into the central Pacific. This bias with respect to observations is most pronounced during the northern warm season, roughly April-November. During December-February (DJF, not shown) all 3 baseline experiments do a reasonable job of simulating precipitation in the tropical Pacific. A NW-SE tilting south Pacific convergence zone (SPCZ) is present south of the Equator between 150°E and 150°W, although in B1 it is less well developed, and more zonally-aligned, than in B2 and B3. During JJA, in the warm pool region (120°E-150°E, Eq-15°N) and also in the central, tropical Pacific (150°E-120°W), precipitation in the experiment with strongest re-evaporation (Exp B3, Fig. 1c) appears closest to the Xie-Arkin climatology. In B1 and B2 the southern Pacific ITCZ appears to grow at the expense of the South Pacific Convergence Zone (SPCZ) and near equatorial, warm pool precipitation.

Total precipitable water (Fig. 2) also changes as re-evaporation strength varies, with the largest global mean value associated with the strongest re-evaporation. In general the experiments with significant re-evaporation B2 (Fig. 2b) and B3 (Fig. 2c) again appear to be in closer agreement with the observational estimate, in this case SSM/I total precipitable water (Alishouse et al., 1990) shown in Figure 2d. Experiment B1 is substantially drier than the observational estimate. Peak values over the Pacific warm pool in B1 are ~ 45 kg m^{-2}, where the observations show values over 55 kg m^{-2}.

Many other aspects of the model's mean seasonal climate also depend on the choice of α_r, including top-of-atmosphere (TOA) radiative fluxes, ocean surface wind stresses, and mean cloudiness (not shown). Some of these changes appear to be related to changes in the model's atmospheric water vapor distribution, and some e.g. wind stress changes, may be related to the large differences in the horizontal distribution of convective heating that occur as α_r varies.

4.2 Precipitation Variance (High-Frequency transients)

High frequency easterly disturbances have been proposed as the origin of off-equatorial rainfall maxima in the atmosphere (e.g. Hess et al. 1993). We will not attempt a detailed analysis of this mechanism in our simulations. However, it is worth establishing the degree to which high frequency transient motions vary with α_r in our model. Figure 3 shows the mean variance for May-November 1990-1991 in precipitation and vertically integrated boundary layer divergence ($\approx\omega_{850}$). Model fields were high-pass filtered with a Lanczos filter (Duchon, 1979) to eliminate modes with time scales longer than 31 d. From Figure 3 it is evident that there have been quantitative changes in the high frequency variability. Both precipitation and ω_{850} variance are higher in Exp B1, in the region of the southern ITCZ (10°S,150°E-120°W). In the SPCZ region (northeast of Australia) and in the northern warm pool region (10°N-20°N, 120°W) variance is higher in Exp B3. This simply reflects the mean distribution of precipitation in each experiment.

Other subtle differences exist between the space-time spectra of high frequency motions in Exp B1 and B3. However, we do not believe these differences are of fundamental importance in producing the changes in precitpitation between Exp B1 and Exp B3. Nevertheless, this issue will be revisited in Section 7.

4.3 Atmospheric water vapor diagnostic fields

During all of the experiments discussed here, a suite of atmospheric water cycle diagnostic fields was saved during model execution. These diagnostic fields included 3D daily averages of the water substance conversion terms in our single condensate phase, prognostic cloud scheme. These are conversion of vapor to cloud condensate, conversion of cloud condensate to precipitation, and finally conversion of precipitation to water vapor, i.e., re-evaporation $\mathcal{R}$. Figure 4 shows the 1990-1991 JJA-average, vertical mass-weighted integral of the rain re-evaporation tendency $<\mathcal{R}>$ in mm d^{-1} from experiment B3 (Fig. 4a), and the proportion of $<\mathcal{R}>$ to the total precipitation generated in the column ϵ_1=$<\mathcal{R}>/\mathcal{P}_g$ (Fig. 4b). We use $\mathcal{P}_g$=$\mathcal{P}_0$+$<\mathcal{R}>$, where $\mathcal{P}_0$ is the precipitation flux at the surface to determine total precipitation generation rate $\mathcal{P}_g$. The figure illustrates the high variability of $<\mathcal{R}>$ and ϵ_1, as well as the surprisingly high values of these quantities. Values for ϵ_1 are typically over 0.5 in the tropics. The global mean value of $<\mathcal{R}>$ for this period is 2.7 mm d^{-1}, which is nearly equal to the globally averaged precipitation rate for the same seasonal average 3.1 mm d^{-1}. By contrast for the same period in B1 (not shown), ϵ_1 is everywhere less than 0.2, and the global mean value of $\mathcal{R}$ is 0.19 mm d^{-1}.

An interesting feature of $<\mathcal{R}>$ in Fig. 4a is its resemblance to rain rate in a double ITCZ regime. It is tempting to conclude from this that the double ITCZ is eliminated by simply evaporating falling rain. This simple view is only partly correct, in the sense that, re-evaporation needs to occur in conjunction with removal of the re-evaporated water vapor from the region of the southern ITCZ. In Section 6 it will be shown that even in experiments with strong re-evaporation, in which dynamical transport of water vapor is suppressed, a southern ITCZ comparable to that in Figs 1a can form.

4.4 Profiles

In order to facilitate comparison of vertical structures, we calculate average profiles in a region bounded by 150W and 170E on the east and west, and by 14S and the Equator in the north-south direction (Box S). As shown in Figure 5, this volume contains most of the spurious southern ITCZ when it forms in experiment B1. We also examine a corresponding volume bounded by the same meridians on the east and west, but by the Equator and 14N in the north-south direction (Box N). These regions will also be used in the next section as control volumes for water vapor budget analyses.

Figure 6a shows profiles of specific humidity in Box S for Exps B1, B2, and B3. Large differences are obvious above the boundary layer, with B1 2 to 3 g kg^{-1} drier than B3 at around 800 mb. The profile from B1 is in the best agreement with NCEP re-analysis profiles (Kalnay et al., 1995) above 700 hPa, but is drier than the analysis below. The profile from B3 is wetter than the re-analysis profile at all levels. The wet bias is most pronounced above 800 hPa. It is distressing that the experiment with the better simulation of precipitation (Exp B3) appears to have larger differences from the NCEP re-analysis. However, it is also possible that these differences are partly model driven. NCEP re-analysis rainfall for example exhibits a stronger double ITCZ than is observed. In situ profiles from the central Pacific (e.g. Yin and Albrecht, 2000) exhibit a good deal of variability depending on location and underlying SST. The mean ITCZ profile from Yin and Abrecht is plotted in Figure 6b against the q-profiles from Box N. This profile is for May-June 1979, and for a longitude range of 90°W TO 120°W. Nevertheless, the Yin-Albrecht profile suggests a dry bias in NCEP re-analysis at midtropospheric levels. The fact that total precipitable water in Exps B2 and B3 is closer to SSMI observations also argues for a dry bias in the re-analysis.

The shape of the water vapor profiles in Exps B2 and B3 is more difficult to explain. The profiles show a pronounced "q-reversal" (Kloesel and Albrecht 1989) between 850 and 900 hPa. However, the height of this feature in observations is typically closer to 800 hPa and is not typical of deep convective regions (Yin and Albrecht, 2000). This structure is likely due to a combination of strong re-evaporation moistening above the MBL, and an excessively shallow, entraining MBL.

Figures 6g,h show profiles of rain re-evaporation $\mathcal{R}$ for JJA 1990-1991 in Box S and N. Moistening from re-evaporation in Exp B3 (thick solid line) peaks immediately above the MBL (900 hPa) with values over 1.5 g kg^{-1} d^{-1}. The structure of $\mathcal{R}$ is at least partially controlled by RH (crosses), with the highest rates occurring in the warm dry air immediately above the boundary layer, and much weaker $\mathcal{R}$ within the moist MBL.

Note that an error in the prognostic cloud formulation allowed very large supersaturation (>300%) to form in cold temperatures (<210 K). This error does not impact the results presented here.

Profiles of moist static energy h and saturated moist static energy h^* for JJA 1990 are shown in Figures 6c,d. The profiles of h for B1 and B3 reflect the large differences in q (Figs. 6a,b). Mid-tropospheric h in B1 is correspondingly lower than in B3, while boundary layer values of h are similar in both experiments. The quantity h^* is of interest primarily because when $(h_{BL} - h^*)>0$, boundary layer parcels are buoyant. The vertical integral of $(h_{BL} - h^*)$ is closely related to CAPE. There are clear systematic differences in the h^* from B1 and B3. The profile from B3 shows greater stabilization at upper levels. The profiles of $(h_{BL} - h^*)$ (not shown) are of relatively small magnitude ($\sim 0.4 \times 10^4$ J kg^{-1} in B3) compared with the vertical variations in h itself ($>1.5 \times 10^5$ J kg^{-1}). In B1 $(h_{BL} - h^*)$ is uniformly smaller than in B3, implying smaller CAPE. This difference arises from the difference in boundary layer q between B1 and B3 which leads to lower h_{BL} in Exp B1.

It is of interest that wherever a direct comparison of thermodynamic profiles for Box S and N was shown, little difference was evident between the box-averaged profiles for a given experiment. This may not be remarkable in the case of Exp B1, where mean precipitation in Box N and S is similar (Table 2), but it is somewhat surprising that Box N and S profiles for Exp B3 were also similar.

5. Water vapor budget analysis

Straightforward analysis of model fields from simulations using different α_r yields a number of interesting features beyond the large difference in southern ITCZ precipitation. Subtle differences in the behavior of high frequency transients were obtained, as well as large differences in atmospheric water vapor content, and other thermodynamic quantities. Unfortunately, none of these differences is easily related to the change in southern ITCZ precipitation with α_r. Analysis of the water vapor budgets in Boxes N and S will at least reveal how the water vapor required to maintain high ITCZ precipitation rates is supplied. Differences in the budgets between Exps B1 and B3 may also suggest mechanisms for suppressing or enhancing precipitation in Box S which contains the spurious, southern ITCZ in Exp B1.

5.1 Box-averaged water vapor budgets

Figure 7 shows a time series of monthly-averaged, area-integrated precipitation $\overline{\mathcal{P}}$ and evaporation $\overline{\mathcal{E}}$ fluxes in these boxes from experiments B1 and B3. A striking aspect of Fig. 7 is the relative lack of structure in the evaporation time series. In both boxes and in both experiments the evaporation remains close to 4.5×10^8 kg s^{-1} for most of the 2 year duration of the experiments. Evaporation in Box N tends to be somewhat higher (0.2–0.5×10^8 kg s^{-1}) than in Box S, and also exhibits a weak annual cycle with peak values of $\sim6.0\times10^8$ kg s^{-1} during the northern cold season. It is of particular interest that the evaporation rates in both experiments are so close. This is despite the fact that boundary layer moisture in the two experiments is different, with box average BL specific humidity in box S typically 1-2 gkg lower in experiment B1 than in B3. Thus, over fixed SSTs, evaporation rates should be higher in B1. However, surface wind speeds in the tropical Pacific are also lower in B1.

In contrast, the precipitation time-series in Fig. 7 show a good deal of variation. Seasonal variations are most pronounced for Exp B3 in Box S (thick, solid line, Fig. 7a). Values range from around 3×10^8 kg s^{-1} from April to September to over 8×10^8 kg s^{-1} during northern winter and late fall (November to March). Integrated precipitation flux for Box S, in Exp B1 (thin, solid line, Fig. 7a) is relatively constant at around 5-6×10^8 kg s^{-1}. The difference between the precipitation time series for B1 and B3 in Box S, reflects the presence of the summertime double ITCZ bias in B1. Comparison with CMAP precipitation estimates for Box S (diamond symbols, Fig. 7a), confirms that in this region, B3 produces a much better simulation of precipitation. Large seasonal excursions in precipitation fluxes are present in the observations, and although the interannual variations in the the November-March maximum are not well captured by B3, the typical amplitude of the seasonal cycle in precipitation (5-6×10^8 kg s^{-1}) is captured. Precipitation in B1 is higher than observed during April-September, consistent with the familiar “double ITCZ” bias pattern. However, during northern winter B1 exhibits a dry bias in Box S. This reflects the relative weakness of the SPCZ in Exp B1.

In Box N (Fig. 7b), the precipitation time series from both experiments are flatter on seasonal time scales. Significant month-to-month fluctuations ($\sim 1\times 10^8$ kg s^{-1}) exist in both experiments. A weak seasonal cycle is evident in the observations, with a distinct minimum in Feb-March. Overall, the precipitation in Box N is relatively constant $\sim 7\times 10^8$ kg s^{-1} throughout the year, for both experiments as well in the CMAP observations.

Time rates of change of the total water vapor mass $\partial_t \langle\overline{\rho q}\rangle$ within boxes S and N are small compared to the fluxes in Fig. 7. Typical values for $< \rho q >$ are $\sim 10^{14}$ kg with month-to-month changes $\sim 10^{13}$ kg. Using 1 month$\approx 3\times 10^6$ s yields $\partial_t \langle\overline{\rho q}\rangle \sim 10^7$ kg s^{-1}. The sum of all fluxes into boxes S and N be similar to this amount. The barely visible bars at the bottom of Fig. 7a, show the implied $\partial_t \langle\overline{\rho q}\rangle$ obtained from the month-to-month changes in total precipitable water within Box S. The large remaining imbalance between precipitation and evaporation, $\overline{\mathcal{P} - \mathcal{E}}$, must be compensated by vertically-integrated advective fluxes of water vapor through the sides of the boxes.

5.2 Advective water vapor transport

The horizontal transport of water vapor in σ-coordinates can be written in flux form:

$$\vec{\nabla} \cdot (\pi \vec{V}_h q)$$

where π is the surface pressure, and V_h is the horizontal wind consisting of zonal component u and a meridional component v. We will examine zonal and meridional water vapor fluxes $\pi u q$ and $\pi v q$. We also examine separately the contributions of $\pi \vec{V}_h \cdot \vec{\nabla} q$ and $q \vec{\nabla} \cdot (\pi \vec{V}_h)$ to the total horizontal advective water vapor tendency in (X). These terms from the daily wind and water vapor outputs on σ-surfaces, along with daily surface pressure output, from our simulations. Some error is unavoidable in this approach since u and v are interpolated from the model's C-grid u and v-points to the grid centers, or p-points, before output. However, below it will be clear that sufficient accuracy has been obtained for the purposes of the analysis here.

The total advective water vapor flux into the boxes above is given by the sum of the vertically integrated fluxes through the four sides;

$$
\begin{aligned}
-\oint_{BoxN,S}\left(\int_0^1 \pi\vec{V}_h q d\sigma\right)\cdot d\vec{A} \quad = \quad & \int_{\text{east}}^{\text{west}} d\lambda \ \int_0^1 a\ cos\phi\ \pi v q\ d\sigma|_{\text{south}} \\
& - \int_{\text{east}}^{\text{west}} d\lambda \ \int_0^1 a\ cos\phi\ \pi v q\ d\sigma|_{\text{north}} \\
& + \int_{\text{south}}^{\text{north}} d\phi \ \int_0^1 a\ \pi u q\ d\sigma|_{\text{west}} \\
& - \int_{\text{south}}^{\text{north}} d\phi \ \int_0^1 a\ \pi u q\ d\sigma|_{\text{east}}
\end{aligned} \tag{5}
$$

where a is the radius of the earth. The four terms on the r.h.s will be denoted by $<\overline{\pi v q}^{\lambda}>^{N,S}_{\text{south}}$, $<\overline{\pi v q}^{\lambda}>^{N,S}_{\text{north}}$, $<\overline{\pi u q}^{\phi}>^{N,S}_{\text{west}}$, and $<\overline{\pi u q}^{\phi}>^{N,S}_{\text{east}}$ for box N or S. The total flux into the box, i.e., the l.h.s of the equation above, will be denoted by $\mathcal{A}_q^{N,S}$. Unless otherwise indicated, calculations of water vapor flux quantities are performed with daily-averaged, model outputs on σ-surfaces. Monthly or seasonal averages of these quantities, therefore include contributions from sub-monthly, transient disturbances.

Monthly-averaged, advective water vapor fluxes for Box S are shown in Figure 8. Positive values for the fluxes in Figure 8 indicate flow of water vapor into the box. First, note the close agreement between the total advective flux $\mathcal{A}_q^S$ and $<\overline{\mathcal{P}-\mathcal{E}}>^S$ confirming that errors in the advective flux calculation are small compared to the quantities of interest. In both experiments, $<\overline{\pi u q}^{\phi}>^{S}_{\text{east}}$ (thin dashed-line) and $<\overline{\pi v q}^{\lambda}>^{S}_{\text{north}}$ (crosses/thin dashed line) remain relatively constant throughout the simulation in both experiments. Water vapor enters Box S through its eastern edge at 150W, consistent with general easterly flow at low levels in the tropical central Pacific. A much weaker outflow of water vapor occurs through the northern edge. The fluxes through the western and southern edges of Box S, $<\overline{\pi u q}^{\phi}>^{S}_{\text{west}}$ (thin solid line) and $<\overline{\pi v q}^{\lambda}>^{S}_{\text{south}}$ (crosses/thin solid line), behave in a more interesting fashion. They are anti-correlated in time, with anomalous inflow of moisture form the west during November-March, coupled with anomalous outflow toward the south. In the northern warm season (April-October) the situation is reversed.

However, $<\overline{\pi u q}^{\phi}>^{S}_{\mathrm{west}}$ and $<\overline{\pi v q}^{\lambda}>^{S}_{\mathrm{south}}$ also exhibit interesting differences between experiments B1 and B3. The seasonal cycles of $<\overline{\pi u q}^{\phi}>^{S}_{\mathrm{west}}$ and $<\overline{\pi v q}^{\lambda}>^{S}_{\mathrm{south}}$ are substantially stronger in B3 (Fig. 8b). Moisture influx from the west during November-March appears to primarily responsible for supplying the anomalously strong precipitation during this period in Exp B3. Westerly moisture influx in B1 (Fig. 8a) is anemic by comparison, with $<\overline{\pi u q}^{\phi}>^{S}_{\mathrm{west}}$ remaining negative for most of simulation. During April-October water vapor exits Box S through the western edge in both experiments at similar rates 2–4$\times 10^{8}$ kg s^{-1}. During this season, which is when double ITCZ bias is strongest, the principal difference between the experiments is in $<\overline{\pi v q}^{\lambda}>^{S}_{\mathrm{south}}$. In Exp B1, there is little net transport of water vapor through the southern edge of Box S. By contrast, in Exp B3 during April-October, there is removal of water vapor ($\sim$2–4$\times 10^{8}$ kg s^{-1}) through the southern edge. Thus, when the spurious double ITCZ is present in Exp B1, the largest difference in the water vapor budget of Box S between Exp B1 and B3, is the fact that water vapor is removed through the southern edge of the box in B3, while little net flux occurs there in B1. This results in year-round advective inflow into Box S in Exp B1. In Exp B3, inflow occurs only during the northern cold season, while net advective outflow occurs in April-October.

The advective fluxes for Box N are shown in Figure 9. Here strong advective inflow ($\sim$2–4$\times 10^{8}$ kg s^{-1}) of water vapor is present most of the year, in both experiments. The zonal fluxes are generally stronger in Exp B3 (Fig. 9b), reflecting stronger zonal trade winds. Also, $<\overline{\pi v q}^{\lambda}>^{S}_{\mathrm{north}}$ exhibits more seasonal variablity in B3, with a distinct period of northerly inflow (June-November) and period of northward outflow (December-April).

5.3 Vertical layering of water vapor transport

It is of interest to know how the differences in water vapor transport between B1 and B3 are distributed in the vertical, and whether are primarily driven by divergent winds ($q\vec{\nabla} \cdot (\pi\vec{V_h})$) or by advection of water vapor horizontal gradients ($\pi\vec{V_h} \cdot \vec{\nabla} q$). Figure 10 shows profiles of horizontal water vapor fluxes into Box S during JJA 1990. The difference $\overline{\pi uq}^{\phi}{}^{S}_{\text{west}} - \overline{\pi uq}^{\phi}{}^{S}_{\text{east}}$ (thin solid line) is the net zonal flux of water vapor into the box, and the difference $\overline{\pi vq}^{\lambda}{}^{S}_{\text{south}} - \overline{\pi vq}^{\lambda}{}^{S}_{\text{north}}$ (thin dotted line) is the net meridional flux into the box. The sum of these is the total horizontal advective flux (thick dashed line) into the box. Below 900 hPa the total flux is near $+1.2\times10^{8}$ kg m^{-2} s^{-1} in both experiments, so that the net transport of water vapor into Box S at low levels appears relatively unaffected by the changes in re-evaporation. The partition of this transport into meridional and zonal components is different, with more meridional transport in Exp B1, consistent with an expected strengthening of the meridional trade winds as precipitation becomes more zonally aligned.

However, above 900 hPa large differences in net water vapor flux are apparent. The net flux in Exp B3 becomes negative implying removal of water vapor from Box S in the free troposphere. This removal of water vapor is accomplished by net meridional transport, which is strongly negative $\sim -1\times10^{8}$ kg m^{-2} s^{-1} in B3. Net meridional transport is also negative in Exp B1, but is much weaker. Further examination of the meridional flux profiles in B3 shows that the free-tropospheric, net meridional flux is dominated by export of water vapor across the southern edge of Box S.

Another interesting feature of the water vapor transport profiles from Exp B3, is the partition between the divergent and gradient components ($q\vec{\nabla} \cdot (\pi\vec{V_h})$ and $\pi\vec{V_h} \cdot \vec{\nabla} q$ resp.). Although the shape of the net flux profile is clearly determined by $q\vec{\nabla} \cdot (\pi\vec{V_h})$ (solid circles) there is a nearly constant and substantial negative contribution from $\pi\vec{V_h} \cdot \vec{\nabla} q$, as indicated by the the difference the total flux profile and the $q\vec{\nabla} \cdot (\pi\vec{V_h})$ profile. In Exp B1 this difference is generally smaller (above 900 hPa) and not systematically negative. Normally the term $q\vec{\nabla} \cdot (\pi\vec{V_h})$ is assumed to overwhelmingly dominate net water vapor transport in the tropics. It appears this assumption is marginal at best in Exp B3.

6. Fictitious drag experiment

The analysis of the simulated water vapor budgets from Exps B1 and B3 suggests that mid-tropospheric water vapor transport plays a key role in suppressing the formation of a spurious, southern ITCZ during the northern warm season. In Exp B3 Meridional transport above 900 hPa removes water vapor from a 1400×4000 km^2 domain (Box S, Fig. 5) including the spurious ITCZ, principally towards the south, and balances the excess of evaporation over precipitation. In Exp B1, for the same domain, precipitation exceeds evaporation with a net dynamical influx of water vapor closing the budget. A similar balance holds for both experiments in an equal-sized domain around the northern ITCZ (Box N, Fig. 5). A naive conclusion that could be drawn from the budget analyses is that reducing meridional transport out of Box S in Exp B3, would lead to an increase in precipitation. This possibility is examined in Exp DM (Table 1), which uses strong re-evaporation as in B3, but incorporates a strong fictitious drag in the meridional direction. The drag acts above 800 hPa, in a band centered on 8°S. The altitiude restriction in the effect of the drag is meant to minimize its impact on low-level, meridional water vapor.

6.1 Basic climate

Mean JJA precipitation for DM is shown in Figure 11a. The precipitation pattern in the tropical Pacific resembles that from B1 more than that from B3. Rates of 8 mm d^{-1} extend well beyond the dateline along 8S. Mean JJA rainfall rates for Boxes S and N from Exps DM, B1 and B3 are given in Table 2. However, other model physics quantities, including precipitation features outside of the tropical Pacific, remain closer to their appearance in B3. Total precipitable water from DM is shown in Figure 12a. Over the tropical Pacific the TPW distribution pattern in Exp DM more closely resembles that from B1, but even there TPW values are more like those in Exp B3. The partition of continental rainfall to oceanic rainfall in the tropics is similar in B3 and DM in JJA and DJF (not shown), suggesting that this aspect of the precipitation simulation is controlled by the re-evaporation.

Figure 13 shows meridional cross-sections of the meridional wind averaged between 170°E and 150°W for JJA. The relatively unremarkable appearance of the flow in Exp DM (Fig. 13c) is noteworthy. We expected the drag in (3,4) to induce a large distortion in the atmospheric circulation. However, the meridional wind in Fig. 13c is quite acceptable, in fact it is closer to that in NCEP re-analysis (Fig. 13d) than the meridional flow in Exp B3 (Fig. 13b). Overall, a routine analysis of model output from Exp DM would not suggest that the physics had been distorted in any way. The large increase in the southward meridional wind, between 30°S and the Equator, for Exp B3, suggests a feedback interaction with increased moist heating in the SPCZ.

6.2 Vertical Profiles

A three-way comparison of vertical profiles, horizontally-averaged in Box S, for JJA 1990, from Exps B1, B3, and DM is shown in Figure 14. Mean profiles of of humidity q (Fig. 14a) and moist static energy h and saturated moist static energy h^* (Fig. 14b) from DM and B3 are nearly identical despite the large difference in mean rainfall for Box S between these two experiments (Table 2). As noted earlier, q qnd h profiles for B1 are quite different from those in B3 above 800 hPa, due to a much drier mid-troposphere in B1. Profiles of h^* from B1 are also distinct from the other two experiments, with less stabilization evident above 600 hPa. Overall, these profiles suggest no obvious connection between the thermodynamic structure of the atmosphere and precipitation in the southern ITCZ. Rather, the mean thermodynamic structure appears to be tied to the choice of α_r in (2).

The mean cloud-base mass flux m_{cb} for different cloud-types (Moorthi and Suarez 1992) in RAS is shown as a as a function of cloud detrainment pressure level in Figure 14c. Below 850 hPa and above 300 hPa all three experiments share similar profiles of m_{cb}. Between 850 and 300 hPa, the m_{cb}-profile for B1 is substantially weaker ($\sim$30-150kg m^{-2} d^{-1}) than those for B3 This implies lower total convective mass flux, and a much higher convective precipitation efficiency ϵ_p in Box S for B1 than for B3 or DM (Table 3). Between 600 and 300 hPa, m_{cb} for DM is stronger by upto 50kg m^{-2} d^{-1}) than that for B3. This accounts for a mean, total convective, cloud-base, mass flux $\sum_l \overline{m_{cb}}$ in DM, that is about 100 kg m^{-2} d^{-1} stronger than in B3 (Table 3). Thus, deep convective clouds are somewhat stronger in Exp DM than in B3.

Although B1 and DM share similar surface precipitation rates, profiles of precipitation fluxes $\mathcal{P}$ (Fig. 14d) are strikingly different. The precipitation profile for B1 increases at a nearly constant rate from 200 hPa all the way to the surface, while that for DM increases rapidly between 200 and 500 hPa, and then remains nearly constant down to the surface. The shape of the $\mathcal{P}$ in B3 is similar, but the generation of precipitation above 500 hPa is weaker than in DM. Profiles of rain re-evaporation $\mathcal{R}$ in B3 and DM (Fig. 14e) are nearly identical. Thus it appears that the re-appearance of the southern ITCZ in Exp DM, is due to an increase in precipitation production above 500 hPa, over that in B3. This increase in precipitation may be related to the increased strength of deep convection in DM.

Also indicated in Figure 14d (thin lines) is the convective component of rainfall for each experiment. This refers to rain produced by RAS, through autoconversion of condensate within convective updrafts. Condensate remaining in the parcel as it detrains is added to the prognostic cloud water scheme, where it autoconverts according to the Sundquist-type formulation in Sud and Walker (1999). Figures 14f and 14g show profiles of cloud condensate production by detraining convection and statistical (RH-based) condensation, as well as, autoconversion of cloud condensate. From Fig. 14f it can be inferred that almost all of the non-convective rain in Box S originates in detraining anvils in all three experiments. The statistical source of cloud water for Exp

DM shows an interesting increase over that in B3. However, integrated over the column it accounts for less than 10% of the increase in precipitation obtained in DM over B3.

In practice, the proportion of convective to total rain in NSIPP-2.0 is controlled by an empirically chosen convective autoconversion rate within RAS (Section 2). The choice of this parameter primarily affects the amount of precipitation originating in anvils vs. convective updrafts, and has little impact on the net rainfall.

Box averaged profiles of horizontal water vapor flux for Box S are shown in Figure 14h. These profiles are consistent with the argument that meridional transport of water vapor plays a key role in eliminating the spurious southern ITCZ. The net transport of water vapor in Exp DM is remarkably close to that from Exp B1, with strong net inflow of water at low levels (below 900 hPa) and weak horizontal water vapor transport above. In Exp B3, by contrast, strong outflow of water vapor above 900 hPa results in a net advective removal of water vapor from Box S despite inflow at low-levels.

7. Analysis of transient motions

The idea that high frequency propagating waves are responsible for the observed characteristics of the ITCZ has been discussed for over 30 years (e.g.; Holton, 1971; Hess et al., 1993; and Gu and Zhang, 2001). While data and models agree that high frequency disturbances (both eastward and westward) are embedded within the ITCZ, proof of causality, i.e., that these motions actually generate the ITCZ in a time-mean sense, has been elusive. In the aquaplanet simulations of Hess et al., zonally asymmetric motion appear to be required to form off-equatorial ITCZs. However, the satellite data analysis of Gu and Zhang suggests that the role of propagating disturbances changes in different sectors of the ITCZ.

As seen in Figure 3 there were different variance patterns for precipitation and boundary layer convergence in B1 and B3. Figure 18 shows variance of these quantities in Exp DM for May-November. A distinct signature of the southern ITCZ is evident in the precipitation variance (Fig. 15a). The magnitude and shape of this variance feature is similar to that in Exp B1 (Fig. 3a). Elsewhere, the precipitation variance for Exp DM is generally between that of B1 and B3, except in the northern warm pool region (120°E-150°W, 10°N-20°N) where it is close to that in Exp B3. Interestingly, the variance of integrated boundary layer convergence ($\approx -\omega_{850}$) in Exp DM (Fig. 15b) is closer overall to that in Exp B3, even in the region of southern ITCZ.

Further differences in space-time power spectra and multiple field, cross-correlations exist. We believe that, while interesting, these differences are not causally responsible for the appearance or disappearance of the spurious southern ITCZ in our simulations. Nevertheless, given the wider interest in the interaction of tropical transients with precipitation, we present a brief analysis of these features in our model.

7.1 Transport

The role of transients in the net water vapor budget for Box S appears to be minor. Figure 16 compares compares the seasonal zonal and meridional water fluxes calculated using daily model output and then averaged in time, with the corresponding fluxes calculated using seasonal means of u, v, q, and π. The difference between these is the net transport across the edges of Box S by sub-seasonal transients. As can be seen in the figure, the profiles are nearly indistinguishable, with the largest differences occurring for the net meridional fluxes in Exp B3 and DM. Here, transient motions lead to a minor $\sim$10% increase in the outflow of water from the box above the PBL, and a similar reduction of the meridional inflow within the PBL.

7.2 Space-Time Frequency Analysis

Two-dimensional FFTs of precipitation were also calculated along 8°N (Figure 17). The figure shows background spectra, i.e., the 2D (ω-k) FFT after repeated applications of a 1-2-1 filter (Wheeler and Kiladis, 1999). The precipitation fields were

31-day high-pass filtered before the FFT was calculated. Four 90 day periods of the filtered data were used; i) May 1, 1990–July 30, 1990; ii) July 31, 1990–October 29, 1990; iii) May 1, 1991–July 30, 1991; and iv) July 31, 1991–October 29, 1991. Power spectra from these four periods were averaged to obtain the results in Figure 17. The results at periods shorter than 31 days (f>0.03 cpd) look qualitatively similar to those shown by Gu and Zhang (2001) for OLR in the ITCZ. The background spectrum is red, with power concentrated in westward moving disturbances (k<0). It is difficult to say whether any of the experiments is qualitatively closer to the Gu and Zhang result. Determination of spectral peaks above the background was not attempted due to the short length of the experiments.

Some interesting differences between experiments are evident. More power in eastward moving disturbances is found in B3 and DM than in B1, suggesting that these modes depend in some way on strong re-evaporation. The speed of the dominant westward moving transients is somewhat slower in B1 than in B3, and there is more power at higher zonal wavenumbers. In this regard, Exp DM appears to lie between B1 and B3. Overall, however the differences in the power spectra of high-frequency transients can be described as subtle.

7.3 Correlation between PBL convergence and precipitation

An intriguing difference between the experiment with weak re-evaporation – Exp B1 and the two with strong re-evaporation – B3 and DM, occurs in the correlation of boundary layer convergence and precipitation. Figure 18 shows maps of the correlation between the two year 1990-91 time-series of daily averages of these quantities at points between 25°S and 25°N. The time-series have been high-pass filtered at 31 days. The correlations for Exp B1 (Fig. 18a) are much higher than for the other two experiments. Peak values above 0.8 are found in the southern ITCZ region, with r>0.7 over broad areas of the tropical Pacific. By contrast, values in B3 and DM are generally between

0.4 and 0.5 in most of the tropics, with small areas of $r>0.6$ over the northern warm pool. This contrast implies stronger coupling between boundary layer dynamics and precipitation in B1 than in B3 or DM. Strong re-evaporative cooling of rain in B3 and DM reduces the net moist heating at low levels compared to that in B1, and therefore also reduces the PBL-top vertical motion associated with precipitation in B3 and DM. Nevertheless, the implications of this coupling for tropical rainfall on climate scales is not clear. Exps B1 and DM exhibit similar seasonal mean precipitation in the tropical Pacific, despite their differences in precipitation-BL coupling

8. PDF Analysis of Convective mass flux and rain generation

The profiles in Figures 14a,b were puzzling in that no large differences in thermodynamic structure are evident between Exps B3 and DM, yet precipitation in the two simulations is quite different. In most of the aspects examined in Section 7, the structure of transients seemed tied to α_r in (2) rather than to the precipitation distribution (e.g.; Figs. 17-18). Thus, the reappearance of the southern ITCZ in Exp DM does not signal a fundamental change in either the thermodynamic state of the simulated atmosphere, or in the nature of simulated high-frequency transients.

Subtle differences are evident in the structure of convective mass flux (Fig. 14c) with Exp DM exhibiting somewhat enhanced deep convection. Profiles of stratiform autoconversion (Fig. 14g) also show more production of rain at upper levels in DM. Thus, the increase in precipitation in DM over B3, is due to a combination of stronger deep convection and more efficient generation of precipitation. Horizontal water vapor fluxes (Fig. 14h) show large differences between B3 and DM, with much weaker, free-tropospheric, transport of water vapor in DM, although the mechanism through which these differences in transport interact with convection and precipitation in Box S is not clear.

Figures 19a-c show joint probably density functions (PDFs) of total precipitation generated $\mathcal{P}_g$ and deep convective mass flux $\sum_{<500} m_{cb}$, defined here as the sum of m_{cb} over all clouds detraining above 500 hPa ($\sum_{<500} m_{cb}$). The PDFs are constructed from the 92 daily averages between June 1, 1990 to Aug 31, 1990 of each quantity at each of 8×17 gridpoints contained within Box S. The total number of points represented by each PDF is thus 8×17×92=12,512. The joint PDFs for all three experiments show a fairly tight functional relationship, with distinct "ridges" oriented along lines sloping between 0.02 and 0.025. The PDF for Exp DM (Fig. 19b) is the least linear, showing both more curvature and more spread than those for B1 (Fig. 19c) and B3 (Fig. 19a). At large values of $\sum_{<500} m_{cb}$ the slope of the precipitation-mass flux relationship increases, indicating more efficient precipitation production by convection. For $\sum_{<500} m_{cb}>1.5$ efficiency in DM appears to be slightly higher than in B3, while for $\sum_{<500} m_{cb}<0.5$ it is slightly lower. Also, for higher mass fluxes there is a wider spread of $\mathcal{P}_g$ values apparent for Exp B3. Certain features of the distributions are difficult to read from the PDFs in Figure 19. One of these is the fact that convection overall is somewhat sparser in B1 than in either B3 or DM, with over 60% of all gridpoint-days free of any convection. In B3 and DM this fraction is around 40%.

From the joint PDFs in Figs. 19a-b we can estimate the increase in Box S precipitation due to increased deep mass flux in DM, with that due to the changing relationship between mass flux and $\mathcal{P}_g$. We apply a least-squares quadratic fit to the $\mathcal{P}_g$ and $\sum_{<500} m_{cb}$ results for Exp B3 (Fig. 19a). These fit parameters are then used with the $\sum_{<500} m_{cb}$ distribution for DM, to estimate the increase in $\mathcal{P}_g$ that results from the simple increase in deep convective mass flux. This calculation shows that increased mass flux accounts for about 70% of $\mathcal{P}_g$ increase obtained in DM. The remaining increase in precipitation generation resuts from the curvature and spread of the PDF in Figure 19b, i.e., from more efficient convective precipitation-generation.

We now use the joint PDFs to identify possible causes for the increased mass flux in DM, as well as, the less important, but significant increase in precipitation-generation efficiency. We isolate 7 populations of comparable size($N{\sim}200$) from the joint distributions. These are indicated by the boxes in Figures 19a-b. The first 5; α^*, α,β, $\alpha^\dagger$, and $\beta^\dagger$ are taken from the area close to the "main ridge" of the joint PDFs, and thus represent the likeliest combination of parameter values. The α's represent conditions with moderate convection in Exps B1 ($\alpha^\dagger$), B3 (α^*) and DM (α), while $\beta^\dagger$ and β represent episodes of intense convection in B1 and DM respectively. Composite humidity profiles for these populations are shown in Figure 19d. Interestingly, all of the profiles from B3 and DM (α^*, α and β) are nearly indistiguishable from one another below 900 hPa, consistent with the weak boundary layer convergence feedbacks implied in Figs. 18b,c. Above 900 hPa, the profile for β (thick dashed line) is wetter by around 1 g kg^{-1}, than those for α and α^*, which remain nearly indistinguishable throughout the column. Thus, a wet, lower, free troposphere appears to coincide with enhanced deep convection in DM. The direction of causality is ambiguous, since convection both entrains air from this layer, and moistens the layer through rain re-evaporation.

An interaction between deep convection and the water vapor profile is present in Exp B1, although more weakly and at somewhat lower levels than in DM. The q-profile for $\beta^\dagger$ (thin dashed curve) is somewhat wetter than that for $\alpha^\dagger$ (thin solid curve) but the difference is largest between 900 and 950 hPa, rather than above 900 hPa. This may be a consequence of the drier free troposphere in B1. Even with wetter than usual conditions, free-tropospheric q above 900 hPa in B1 is still not high enough to provide a significant boost to deep convection, while anomalously high q in the moister 900-950 hPa layer can. The shape of the profile with moistening at 900-950 hPa suggests deepening of the boundary layer. This would be consistent with the strong convergence/precipitation interaction implied by Figure 18a.

Next we examine the composite profiles for γ and δ (Fig. 19e). These profiles represent episodes of intense convection but varying amounts of precipitation generation. Relatively inefficient precipitation is represented by γ (thick dashed line), while efficient precipitation is represented by δ (thick solid line). Increased precipitation efficiency appears also to be associated with enhanced humidity above 900 hPa. Interestingly, the boundary layer humidity in γ is slightly (but significantly) higher than in δ, suggesting that enhanced boundary layer moisture is required to obtain large convective mass fluxes with a dry free-troposphere.

The preceeding analysis suggests that both convective mass flux and precipitation efficiency increase with enhanced free-tropospheric humidity. Still, the average humidity profiles for Box S (Fig. 14a) show that, if anything, Exp DM is slightly drier between 900 and 600 hPa than B3. Thus, the simple mean does not explain the increased precipitation in DM. Figures 19f-g show PDFs for Box S of $<q>^*_{800-900}$, the vertical average of daily q between 800 and 900 hPa. The total PDFs (Fig. 19f) show that the humidity distribution in Exp DM (dashed line) is somewhat bi-modal, with a dry mode near 8g kg^{-1} and a wet mode near 12 g kg^{-1}. While the overall Box S mean for DM is lower by around 0.5 g kg^{-1} than that for B3 (vertical lines at bottom of figure), the maximum values of $<q>^*_{800-900}$ reached in DM are higher.

The PDFs of $<q>^*_{800-900}$ in Fig. 19f include both convective and non-convective situations. We use a criterion of $\sum_{<500} m_{cb}$>0.5kg m^{-2} d^{-1} to obtain convecting sub-populations of approximately equal size in Exps B3 and DM. This criterion yields a convecting sub-population containing around 42% of the total population in B3, and around 44% in DM. However, in terms of $\mathcal{P}_g$ these sub-populations account for 79% of the total in B3 and 90% in DM. PDFs of $<q>^*_{800-900}$ for these convecting populations and the complementary weakly-convecting (quiescent) populations are shown in Figure 19g. For Exp DM the PDFs are nearly disjoint explaining how high $<q>^*_{800-900}$ can enhance precipitation in Box S without raising the seasonal, Box S mean value. This is a simple consequence of the relative concentration of areas with strong convection. Interactions between $<q>^*_{800-900}$ and convection occur in a restricted area, while outside of convective regions $<q>^*_{800-900}$ is controlled by other factors. We speculate that the increased separation between the convecting and quiescent PDFs (and the bimodality of the total PDF) for Exp DM arises from reduced dynamical transport of water vapor away from convecting regions.

9. Summary and Discussion

We examined the sensitivity of simulated, tropical Pacific rainfall patterns in the NSIPP-2.0 AGCM to the strength of parameterized rain re-evaporation (Eqs. 1,2). With weak re-evaporation (Exp B1, Table 1) the model develops a pronounced "double ITCZ" bias during the northern warm season, which maximizes during July-August (Fig. 1a). As re-evaporation increases the spurious southern ITCZ decreases in strength (Figs. 1b,c), and the overall precipitation simulation over tropical and subtropical oceans improves. Analysis of moisture tendencies shows that in the high re-evaporation experiment (Exp B3), the column integrated re-evaporation tendency is frequently as large as, or larger than, the precipitation at the surface (Fig. 4) implying that half or more of all the precipitating water substance generated, evaporates on its way to the surface.

Two control regions containing the northern and southern ITCZs (Box N and Box S, Fig. 5) were selected for closer analysis of Exps B1 and B3, the simulations with the weakest (B1) and strongest (B3) re-evaporation. Average profiles of q, h, and h^* (Fig. 6) showed large inter-experiment differences, but little difference between northern and southern control volumes. This was puzzling in the case of Exp B3, since precipitation rates in Box S and Box N are quite different (Table 2).

Water vapor budgets were calculated for each of the control regions. A striking aspect of these budgets was the lack of variation in total surface evaporation (Fig. 7). Total surface evaporation was similar in both control regions, as well as, in both experiments. The budgets also showed that during the northern warm season (April-October) evporation exceeds precipitation in Box S in Exp B3 (Fig. 7a), and thus water vapor must be advectively exported from this region. In Exp B1 precipitation exceeded evaporation during the same period, which coincides with the period of pronounced double ITCZ bias. By contrast the northern control region, Box N, required advective import of water vapor in both experiments (Fig. 7b).

Analysis of horizontal water vapor fluxes through the sides of each box (Figs. 8 and 9) revealed an inter-experiment difference in the sign of meridional water vapor fluxes through the southern edge of Box S, during northern warm season. In Exp B1, there is weak, but consistent, inflow of water vapor into Box S from the south, when the spurious ITCZ is present. By contrast, in Exp B3 there is significant outflow during the same period. Profiles of horizontal water vapor fluxes (Fig. 10) revealed that the largest inter-experiment differences in meridional water vapor flux were found above 900 hPa. Below 900 hPa, there is strong meridional inflow of water vapor into Box S, in both experiments. However, above 900 hPa, Exp B3 exhibits a layer of strong advective water vapor removal from Box S, while Exp B1 exhibits weak transport in the same layer.

These budget analyses suggested that meridional water vapor transport may play a key role in the dynamics of the southern ITCZ. In order to explore this possiblity we performed an idealized experiment with strong re-evaporation (Exp DM, Table 1), in which a fictitious drag (Eqs. 3,4) was added to the meridional momentum equation. The JJA precipitation pattern over the tropical Pacific in DM (Fig. 11a), closely resembles that from B1 - the experiment with weak re-evaporation. This is true despite the fact that other quantities in DM related to moist physics resemble those in B3, including total precipitable water (Fig. 12), and humidity and static energy profiles (Figs. 14a-b).

However, profiles of advective water vapor transports are again similar in B1 and DM (Fig. 14h). Thus, while the strength of re-evaporation controls the gross thermodynamic structure of atmosphere in these three experiments, precipitation in the southern ITCZ responds to dynamical factors as well.

The close similarity of q and h profiles in DM and B3 (Figs. 14a,b) raises the obvious question of how the precipitation patterns can be as different as they are in these simulations, without attendant changes in thermodynamic structure of the atmosphere. Straightforward analysis of high frequency transients (Figs. 15-18) does not suggest a clear answer. An analysis using probability density functions (PDFs) of convective mass flux, precipitation, and humidity (Fig. 19) is more promising. This analysis shows that enhanced humidity above 900 hPa is associated both with stronger deep convective mass flux, and with more efficient precipitation generation for a given mass flux in Exp DM. PDFs of humidity in the 900-800 hPa layer, over the southern ITCZ, show that DM possesses a distinctly bimodal water vapor distribution (Fig. 19f,g) with a wet, convecting mode and a dry, quiescent mode. Exp B3, on the other hand, exhibits a nearly normal PDF for $<q>^*_{800-900}$ with reduced spread. Convecting and quiescent modes in $<q>^*_{800-900}$ are present, but are less separated. The enhanced precipitation in DM is associated with the high $<q>^*_{800-900}$ values in the convecting mode, while the overall mean $<q>^*_{800-900}$ in DM remains similar to that in B3, because of the low values in the quiescent mode.

Bi-modal tracer PDFs are characteristic of flows in which mixing is inhibited (e.g.; Sparling, 2000 ; Zhang et al., 2003) We speculate that the weak dynamical transport in Exp DM is inefficient at eliminating water vapor anomalies above the boundary layer. This allows free-tropospheric humidity to build-up in convecting regions, as water vapor is added by re-evaporating rainfall. Free-tropospheric water vapor is available to entraining convective plumes and enhances both convective mass flux, and precipitation generation. This feedback is interrupted by strong advective transports in Exp B3. In Exp B1, a more standard CISK-type feedback between convection and boundary layer convergence (Fig. 18) may be operating. We believe these two forms of convection/precipitation feedback are analogous, with the crucial difference that the boundary layer convergence feedback cannot be interrupted by advective removal of water vapor in the free-troposphere. It is noteworthy that boundary layer moisture in Exp DM appeared to have only a minor role in separating moderate and intense convection epsiodes (Fig. 19d-e).

The main results of this study can be summarized as follows: 1) Rain re-evaporation has a major impact on tropical precipitation in the NSIPP AGCM with stronger re-evaporation leading to less double ITCZ bias. 2) The double ITCZ bias appears to be suppressed by a combination of; a) free-tropospheric moistening and advective removal from the southern ITCZ domain, and b) weakened coupling between precipitation and boundary layer convergence. The sensitivity of precipitation in the NSIPP model to re-evaporation may depend on both factors. Exp DM shows that that changes in the meridional circulation response to re-arrangements of precipitation can eliminate the sensitivity. It is also possible that if boundary layer feedbacks remained stronger in simulations with high re-evaporation, the double ITCZ bias would persist in the simulations.

This study did not address the fundamental mechanisms responsible for the placement of the ITCZs. The similarity in rainfall patterns in B1 and DM, resulting from what appear to be different feedback mechanisms, suggests that the initial organization of precipitation is similar in the experiments. The apparent lack of boundary layer feedbacks in DM, would argue against theories for this initial organization that rely on wind-driven evaporation effects (e.g. Chao 2000). Wave driven theories (e.g.; Holton, Hess et al.) are less contradictory with our results, since the wave motions that organize precipitation could act in a deeper layer of the atmosphere. The strong southward flow of water vapor away from the southern ITCZ in B3 appears to be associated with convection in the SPCZ. We speculate that precipitation in the tropical south Pacific will fall into one of two regimes - a double ITCZ regime, or an SPCZ regime, depending on the ability of the SPCZ to interrupt the strong feedbacks between convection and atmospheric humidity that are favored at ITCZ latitudes.

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